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A ten-year water balance of a mountainous semi-arid watershed

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Abstract

Quantifying water balance components, which is particularly challenging in snow-fed, semi-arid regions, is crucial to understanding the basic hydrology of a watershed. In this study, a water balance was computed using 10 years of data collected at the Upper Sheep Creek Watershed, a 26-ha semi-arid mountainous sub-basin within the Reynolds Creek Experimental Watershed in southwest Idaho, USA. The approach computed a partial water balance for each of three landscape units and then computed an aggregated water balance for the watershed. Runoff and change in ground water storage were not distinguishable between landscape units. Precipitation, which occurs predominantly as snow, was measured within each landscape unit directly and adjusted for drifting. Spatial variability of effective precipitation was shown to be greater during years with higher precipitation. Evapotranspiration, which accounted for nearly 90% of the effective precipitation, was estimated using the Simultaneous Heat and Water (SHAW) Model and validated with measurements from Bowen ratio instruments. Runoff from the watershed was correlated to precipitation above a critical threshold of approximately 450 mm of precipitation necessary to generate runoff ($r^2 = 0.52$). The average water balance error was 46 mm, or approximately 10% of the estimated effective precipitation for the ten-year period. The error was largely attributed to deep percolation losses through fractures in the basalt underlying the watershed. Simulated percolation of the water beyond the root zone correlated extremely well with measured runoff ($r^2 = 0.90$), which is derived almost entirely from subsurface flow. Above a threshold of 50 mm, approximately 67% of the water percolating beyond the root zone produces runoff. The remainder was assumed to be lost to deep percolation through the basalt. This can have important ramifications in addressing subsurface flow and losses when applying a snowmelt runoff model to simulate runoff and hydrologic processes in the watershed.

Keywords: Precipitation; Snow; Hydrologic balance; Semi-arid rangeland; Evapotranspiration

1. Introduction

Quantifying the components of the water balance for a watershed is crucial toward understanding of the dominant hydrologic processes occurring in a basin. Although a number of water balance studies have been conducted for a variety of watersheds throughout the world (e.g. Scanlin, 1994; Yin and Brook, 1992; Kattelmann and Elder, 1991; Motoyama et al., 1986; Mather, 1979; Clarke and Newson, 1978; Davis, 1971), the water balance of snow-fed, semi-arid, rangeland watersheds presents some interesting challenges. These watersheds, dominated by precipitation and evaporation, exhibit a high degree of variability in snow distribution and vegetation communities on scales much smaller than that addressed by most hydrologic modeling. Thus, these basins pose a unique set of problems for hydrologists which include: drifting snow; a water balance dominated

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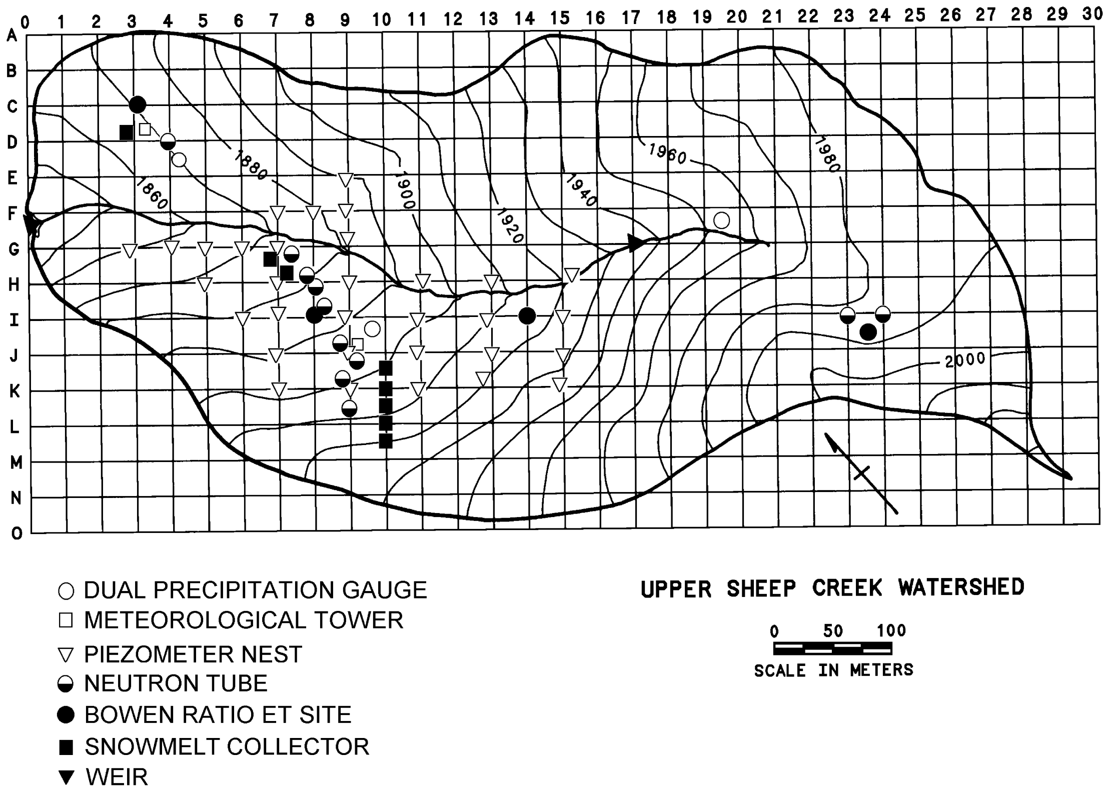

Fig. 1. Topography and instrument locations within the Upper Sheep Creek Watershed.

by evapotranspiration (ET); spatially varying plant communities associated with changes in soil and effective precipitation; the soil water decoupled from ground water after snowmelt; and intermittent streamflow which lasts only a few months in the early spring. Accounting for the variability in effective precipitation and estimating ET from plants which are water-stressed for much of the year make an accurate accounting of the water balance a difficult task.

Flerchinger et al. (1998a) compared two approaches of computing a water balance using 2 years of data from the 26-ha Upper Sheep Creek Watershed, a snow-dominated, semi-arid, rangeland watershed located in the Reynolds Creek Experimental Watershed in southwest Idaho, USA. They found that the difference in ET estimates assuming uniform vegetation compared to disaggregating the watershed into vegetation zones was much greater during a dry year when water was limiting. This suggested that the spatial variability of plant communities and leaf area index is less critical under conditions with ample water than for semi-arid or arid environments.

As a follow-up to the work of Flerchinger et al. (1998a), this study computes an annual water balance of the Upper Sheep Creek Watershed for each of 10 years (1985 through 1994). The approach used was to disaggregate the watershed into landscape units, similar to the approach of Flerchinger et al. (1998a). The objectives of a more thorough analysis of this unique data set are: (1) to better quantify the source of errors in the water balance; and (2) to gain a better understanding of the hydrologic processes occurring in the watershed than could be gained by the two non-contiguous years studied previously. By using 10 continuous years of data, a better accounting of carryover from year to year and a more rigorous analysis of precipitation input were achieved.

2. Description of the study site

The Upper Sheep Creek Watershed is a 26-ha mountainous rangeland watershed located within the Reynolds Creek Experimental Watershed in the Owyhee Mountains of southwest Idaho, USA. A detailed study of the Upper Sheep Creek Watershed was initiated by the USDA-ARS Northwest Watershed Research Center in 1984 and expanded in 1989. Annual precipitation is approximately 508 mm, most of which is snow. Numerous investigations have been conducted to define the geology of the watershed (Winkelmaier, 1987; Mock, 1988; Stevens, 1991) and to better understand the processes controlling the hydrologic response of this mountainous watershed (Cooley, 1988; Duffy et al., 1991; Flerchinger et al., 1992; 1993, 1994, and 1996; Deng et al., 1994; Unnikrishna et al., 1995; Neale et al., 1995; Tarboton et al., 1995; Luce et al., 1998). Locations on the watershed are referenced by a grid system as illustrated in Fig. 1.

The site has considerable spatial variability in soils, vegetation and snow cover. Prevailing southwesterly winter storms cause deep drifts to form near the crest of the leeward northeast-facing slopes where tall shrubs and aspen thickets are found. Vegetation on the wind-swept, west-facing slope is very sparse, where large areas are bare of snow for much of the winter. By contrast, snow cover in the large drifts typically remain into June. Three distinct vegetation types can be identified on the Upper Sheep Creek Watershed: low sagebrush, mountain big sagebrush and aspen. Low sagebrush areas are located predominantly on the west-facing slopes and are bare of snow for much of the winter. North-facing slopes are covered with predominantly mountain big sagebrush and typically accumulate about a meter of snow during the winter. Aspen thickets are established on the upper portions of the northfacing slopes where large snow drifts form annually.

Spring snowmelt is the primary source of runoff from the basin and provides the driving hydrologic force for runoff and subsurface flow. Nearly all water reaching the stream is subsurface flow; overland flow is seldom observed in the basin. The geology of Upper Sheep Creek consists of variably fractured and altered basalt underlain by a thick dense basalt at a depth of 20–30 m. Geophysical studies of the area indicate that the surface of the dense basalt closely follows surface topography (Winkelmaier, 1987; Mock, 1988; Stevens, 1991) so that the watershed boundary for the ground water flow is approximately the same as for the surface.

3. Field data

The instrumentation network at Upper Sheep Creek was constructed in 1984 and is described in detail by Flerchinger et al. (1998a). Hourly meteorological data were collected on the north-facing slope near J9 (Fig. 1) and on the west-facing slope near D4. Precipitation was measured at locations D4, I10 and F19 using the dual-gauge system especially designed for the wind and snow conditions prevalent in the area (Hamon, 1973; Hanson, 1989). A Bowen ratio unit to estimate ET and the surface energy balance was rotated at weekly intervals between the three vegetation types when vegetation was actively transpiring during 1990. Three separate units were operated continuously at sites C4, I8, and I 24 during the 1993 growing season. As part of a collaborative study, Artan (1996) also operated a Bowen ratio unit at I14 for part of the 1993 season. Three sets of melt collectors were used to monitor snowmelt rate from different snow accumulation areas. A transect of 5 melt collectors was located beneath the large drift (K10–M10), a total of six collectors was located in the mountain big sagebrush area downslope of the drift (near H7), and two collectors were located on the windswept slope within the low sagebrush unit (near D3). Streamflow was monitored continuously at the outlet of the watershed using a permanent v-notch weir placed within a concrete cutoff wall extending down to bedrock. Soil water measurements were collected from 1990 to 1993 using a neutron probe approximately every two weeks after the snow had melted at 11 profiles ranging in depth from 100 to 260 cm. A total of 53 piezometers were located at 32 locations within the study area and vary in depth from 4 to 22 m. Water level in the piezometers was recorded hourly using pressure transducers.

Vegetation data collected periodically at Upper Sheep Creek since 1984 indicate that the leaf area index of the low sagebrush, mountain big sagebrush, and understory of the aspen site at the peak of the

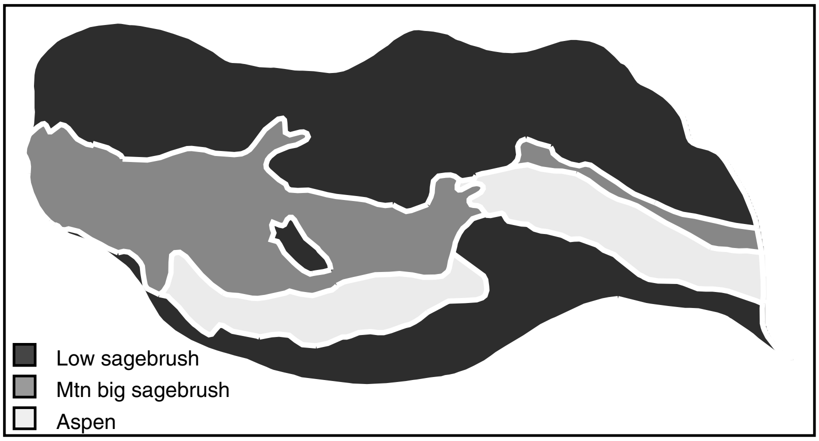

Fig. 2. Disaggregation of the Upper Sheep Creek Watershed into three zones based on vegetation and snow accumulation. The low sagebrush, mountain big sagebrush and aspen zones comprise 58.9%, 26.6% and 14.5% of the watershed, respectively.

growing season is approximately 0.4, 1.2 and 1.0 based on point frame measurements. Measurements within a vegetation type were typically taken with a 20-point frame at 6 locations on each of four 30-m transects to obtain a spatial average of an area. Transect measurements and destructive sampling of a single representative aspen, yielded a tree leaf area index for the aspen site of 2.04.

Additional field data collected throughout the winter season included snow depth and water equivalent at each grid point on the watershed (Fig. 1) using standard snow sampling techniques and the Rosen-type snow tube. Snow measurements were taken at approximately two week intervals during the later part of the snow accumulation period and throughout the snow melt period (March through May).

4. Estimation of water balance components

Water years (October 1 through September 30) 1985 through 1994 were used for analysis of the water balance. The water balance of the watershed may be expressed as:

$$P - \mathrm{ET} - R - \Delta S - \mathrm{DP} = 0 \tag{1}$$

where P is the precipitation, ET the evapotranspiration, R runoff, ΔS the change in water storage, and DP the deep percolation. Some of the underlying assumptions used in our analysis include: the dense basalt underlying the watershed is impermeable and allowed no deep percolation from the watershed to occur (DP = 0); all outflow from the basin was through the outlet weir, and uniform vegetation conditions within each landscape unit.

Because of the heterogeneity of the watershed, it was broken into three areas based on similarity in soils, vegetation, and snow accumulation. A partial water budget was computed for each of the three landscape units identified by Flerchinger et al. (1998a) and delineated in Fig. 2. These are referred to as low sagebrush, mountain big sagebrush, and aspen, which comprise 58.9, 26.6 and 14.5% of the watershed, respectively. A detailed description of the different zones is given by Flerchinger et al. (1998a) and summarized below.

The low sagebrush zone (*Artemisia arbuscula*) has sparse vegetation with some grasses (*Poa secunda*) and considerable bare ground. Soils are generally shallow (30 cm) to basalt bedrock, have high rock content (>50%), relatively high clay content ($\approx$ 25%) argillic horizons and thin (<10 cm) silt loam surface horizons. Snow cover is relatively shallow for most of the winter (<60 cm).

The mountain big sagebrush zone (*Artemisia tridentata vaseyana*) is completely covered with taller sagebrush, snowberry (*Symphoricarpos spp.*) and grasses. Soils are deeper to basalt bedrock (>100cm), have much lower rock contents within the upper meter, little argillic horizon development

Table 1
Initial and final drift factors applied to winter periods for each zone (initial drift factor computed prior to accounting for sublimation losses)

Water year	Low sagebrush		Mountain big sagebrush		Aspen	
	Initial drift factor	Final drift factor	Initial drift factor	Final drift factor	Initial drift factor	Final drift factor
1985	0.88	1.06	0.88	0.99	2.22	2.30
1986	0.93	0.97	0.90	0.98	3.17	3.20
1987	1.07	1.22	1.03	1.18	1.70	1.97
1988	0.87	1.14	0.81	1.00	1.93	2.21
1989	0.92	1.19	0.92	1.04	2.37	2.46
1990	0.80	0.89	0.62	0.68	2.34	2.42
1991	0.72	0.78	0.82	0.92	1.65	1.87
1992	0.74	0.92	0.69	0.79	2.07	2.31
1993	0.77	0.82	0.90	0.96	1.85	1.88
1994	0.66	0.80	0.69	0.80	2.08	2.23

and a relatively thick (50–100 cm) silt loam horizon. Winter snow accumulation in this unit is typically 1 m.

The aspen zone consists of a thick stand of aspen (*Populus tremuloides*)and willow (*Salix spp.*). Soils are very deep to bedrock (>200 cm), virtually rock free, have little argillic development and are almost entirely composed of silt loam material. This unit is characterized by areas where large snow drifts form annually and persist after snow has disappeared from the remainder of the watershed. Winter snow depth varies from 1 m to more than 8 m.

5. Precipitation

To account for snow drifting into the watershed, precipitation measurements during the snow covered period were compared with measurements from the snowmelt collectors and snow measurements taken at the grid points within each zone. Based on these comparisons, a drifting factor was applied during the snow covered period to the precipitation (rain + snow) measured at D4 for the low sagebrush site and at I10 for the mountain big sagebrush and aspen zones to compute an effective precipitation. Using these preliminary adjustment factors, effective precipitation estimates were made and the SHAW model (discussed subsequently) was run to estimate the amount of snowcover lost to sublimation. The drift factors were then re-computed to account for sublimation. A summary of the drift factors for each zone is given in Table 1. The drift factors for 1990 and 1993 in Table 1 differ somewhat from those reported by Flerchinger et al. (1998a) due to the fact the drift factors were computed previously using the period from November to March (November to April for the aspen site); in the current study, the procedure was refined to consider only the snow covered period. However, the difference in average effective precipitation for the watershed differed from the previous study by only 3 mm for 1990 and 11 mm for 1993.

Inspection of drift factors for the sagebrush areas in Table 1 indicate that these areas are scour zones during some years (drift factor <1.0) and deposition zones during other years. This is due to the fact that vegetation in these areas have a limited capacity to store snow, after which the snow tends to be blown away. During years with low snow accumulation or transient snow conditions, the vegetation captures the snow blowing across the area and acts as a deposition zone as long as the snow depth does not exceed the vegetation height. When the snow depth exceeds the vegetation height, additional snow is blown off the area, and the area is a scour zone.

6. Evapotranspiration

ET was estimated using model simulations from the Simultaneous Heat and Water (SHAW) Model (Flerchinger et al., 1996; Flerchinger and Pierson, 1991). The model simulates a one-dimensional profile with provisions for a plant canopy, snow, residue and

Table 2
Comparison of measured ET with single-year simulations reported by Flerchinger et al. (1998a) and continuous multi-year simulations used in the current study

Year	Days measured	Measured ET (mm)	Single-year simulation (mm)	Multi-year simulation (mm)
Low sagebrush				
1990	24	41	44	45
1993	73	145	140	137
Mountain big sagebrush				
1990	27	74	67	56
1993	86	279	273	279
Aspen				
1990	23	85	89	88
1993	48	196	206	210

a soil profile. It has been applied and tested extensively in the Upper Sheep Creek Watershed and the surrounding Reynolds Creek Experimental Watershed (e.g. Flerchinger and Pierson, 1991; Flerchinger et al., 1994, 1996), as well as to a diverse range of sites including the desert southwestern United States (Flerchinger et al., 1998b; Dixon, 1999) and the northern latitudes of Canada and Alaska (Flerchinger et al., 1990; Hayhoe, 1994; Sharratt and Flerchinger, 1995). The model is capable of simulating the surface energy balance and ET from a multi-species plant canopy (including standing dead plant material) using detailed physics of heat and water transfer through the soil–plant–atmosphere continuum.

Flerchinger et al. (1998a) compared simulations with measurements from the Bowen ratio units near grid points C4, I8 and I24 within the Upper Sheep Creek Watershed. Single-year simulations using a 4-m soil profile for each of the landscape units were conducted for 1990 and 1993 and compared with measured data for periods where energy balance data were available. In the current study, a continuous simulation was conducted from October 1983 to September 1994. The initial year was used as a startup year to allow the model to adjust to the climatic conditions since initial soil water measurements were not available. Measured and simulated values are given in Table 2.

ET values reported herein (Table 2) differ slightly from those reported by Flerchinger et al. (1998a) due to a combination of differing drift factors and the multi-year model simulations used in the present study. The largest difference is for the 1990 mountain big sagebrush site, where the annual ET estimate dropped by 55 mm. The change in the drift factor for this year resulted in a 27 mm decline in estimated effective precipitation for the mountain big sagebrush. Even so, simulated ET exceeded precipitation on the mountain big sagebrush area during 1990, resulting in no contribution to runoff or ground water flow. Errors in the initial soil water content tended to affect the mountain big sagebrush area more because the root-zone water holding capacity of both the low sagebrush with its shallow rooting depth and the aspen area with its large effective precipitation tends to be satisfied during most years. The rooting depth of the low sagebrush site, however, was increased to 50 cm for these simulations to yield late summer ET rates that better compared with measurements. Changes in estimated ET from these modifications occurred exclusively in the late summer when water in the soil profile became limiting. Even with these changes, basin-averaged ET estimates for 1990 and 1993 were within 8 and 1 mm of estimates conducted by Flerchinger et al. (1998a).

7. Change in storage

Change in storage within the watershed was separated into soil water storage and ground water storage. Changes in soil water stored within each landscape unit were determined from soil water content measurements and model simulations. Changes in ground water storage for the entire watershed was estimated from ground-water levels measured in piezometers.

Table 3
Summary of water balance for each landscape area (in mm)

Water year	Low sagebrush				Mountain big sagebrush				Aspen			
	Precip.[a]	ET[b]	ΔS (soil)	Areal contribution[c]	Precip.[a]	ET[b]	ΔS (soil)	Areal contribution[c]	Precip.[a]	ET[b]	ΔS (soil)	Areal contribution[c]
1985	489	396	−39[b]	132	557	572	−36[b]	21	1082	446	14[b]	622
1986	482	378	4[b]	100	614	580	2[b]	32	1284	491	−23[b]	816
1987	406	444	−33[b]	−5	434	608	−141[b]	−33	601	568	−53[b]	86
1988	296	335	−38[b]	−1	339	408	−54[b]	−15	541	516	−5[b]	30
1989	417	389	31[b]	−3	584	531	62[b]	−9	1159	522	23[b]	614
1990	390	412	−21[b]	−1	441	494	−44[b]	−9	907	527	24[b]	356
1991	296	298	3	−5	415	418	−44	−9	588	523	58	7
1992	234	243	19	−28	272	307	−26	−9	469	509	−157	117
1993	516	364	37	116	777	510	162	105	1268	437	240	591
1994	237	347	−110[b]	0	306	489	−171[b]	−12	586	498	−55[b]	143
Average	376	360	−15	31	474	492	−24	6	849	504	7	338

[a] Precipitation adjusted for drifting of snow.
[b] Indicates model-simulated values.
[c] Areal contribution to runoff and ground water recharge.

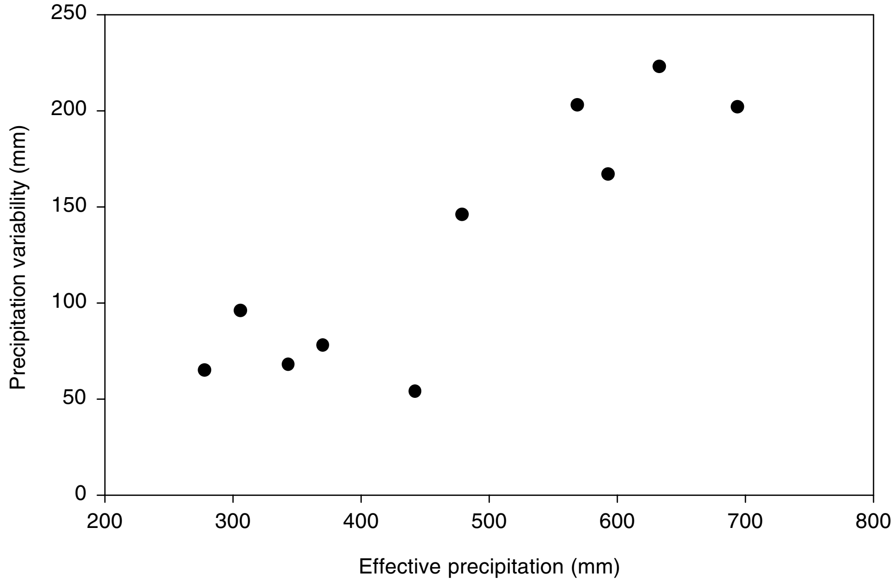

Fig. 3. Variability in effective precipitation between landscapes (defined as the root mean square deviation in annual precipitation) versus annual effective precipitation.

Soil water content measurements were available from 1990 to 1993. For these years, change in storage within the soil profile for each landscape unit was based on soil water profiles measured near October 1 using neutron access tubes. Tubes located at D4 (100 cm deep) and I8 (260 cm deep) were found to be representative of the low sagebrush and mountain big sagebrush areas, while tubes at I23 and I24 (200 cm deep) were used for the aspen area. Variation in total soil water content between locations within the mountain big sagebrush and aspen areas, respectively, was within $\pm 5\%$ for a given year. (Only one soil moisture tube existed in the low sagebrush area.)

Based on the measurements, change in soil water storage for 1991, 1992 and 1993 were $+5$, -18, and $+100$ mm. Two approaches were evaluated for estimating soil moisture storage for years when soil water profile measurements were not available. One approach used the change in storage simulated by the SHAW model for each landscape unit. For the second approach, a regression analysis was conducted to relate soil water storage for the three landscape areas to soil moisture profiles measured at Lower Sheep Creek located approximately 2 km away. The SHAW model simulated a change in soil water storage of -1, -20, and $+140$ mm for years 1991 through 1993, whereas the regression analysis yielded a change of -5, 2, and 4 mm. Based on this comparison, the changes in soil water simulated by the SHAW model were better estimates than those given by the regression analysis, and therefore were used for years when soil water measurements were not available.

Piezometer measurements of groundwater levels were used to estimate the change in ground water storage for each year. Most of the piezometers were located in the mountain big sagebrush area because ground water is most active in this area (Figs. 1 and 2). Large quantities of melt water from the drift along grid lines K and L percolate downward to the dense basalt layer and flow laterally downslope to the stream (Deng et al., 1994). Based on the water level measurements and geophysical interpretation of the bedrock topography, the saturated ground water zone surrounding the creek was estimated for each year, and a weighted average of the change in water level was computed. The change in ground water storage was then estimated using a porosity of 10%.

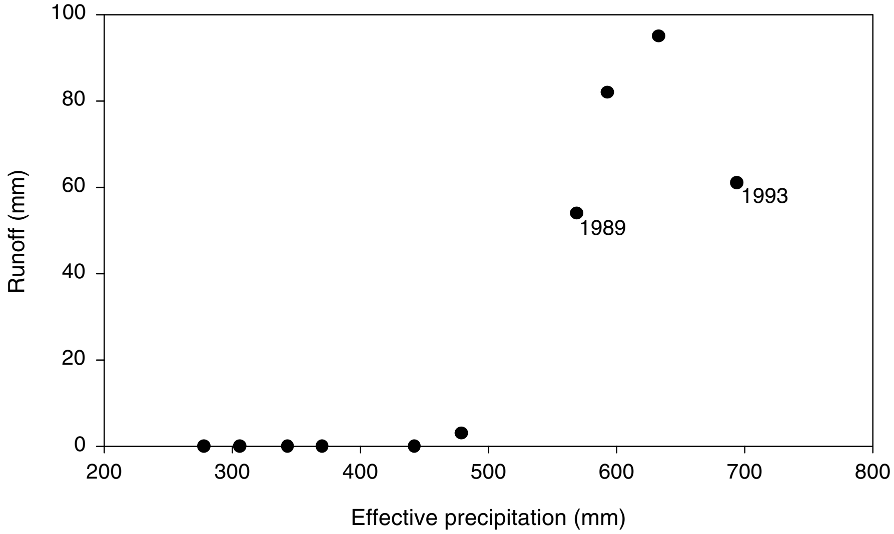

Fig. 4. Measured runoff versus average areal precipitation on the Upper Sheep Creek Watershed.

8. Results and discussion

Precipitation, particularly snow, can vary considerably within a relatively small watershed such as Upper Sheep Creek due to areas of deposition (aside from drifting of snow) associated with local topography (Hanson, 1982). Measured average annual precipitation over the ten-year period was 382, 494, and 486 mm for sites D4, I10 and F19, respectively. Much of this variability occurred during the snow accumulation period from November through March. Measured precipitation during this period averaged 229, 316, and 312 mm, respectively. However, this is only a fraction of the variability in effective precipitation that resulted from drifting. Average annual effective precipitation between landscape areas varied from 376 to 849 mm, as summarized in Table 3. Consequently, it is imperative to account for the variability in precipitation and snow redistribution in addressing the hydrologic processes occurring in the watershed.

Variability in effective precipitation between landscapes (defined as the root mean square deviation in annual effective precipitation for a given year) increases with effective precipitation, as shown in Fig. 3. This is largely due to greater accumulation of snow in the aspen zone during years with higher winter precipitation. The size of the snow drift in the aspen zone can vary considerably from year to year, whereas peak snow accumulation in the low sagebrush and mountain big sagebrush areas varies relatively little.

The relation between runoff and effective precipitation for the watershed is plotted in Fig. 4. Based on these observations, it is apparent that a threshold in precipitation must be reached before runoff occurs. This is fairly typical of the ephemeral snow-dominated streams in the area. Above the threshold in precipitation, the relation between runoff and effective precipitation for Upper Sheep Creek is fairly linear. The notable outlier occurred in 1993, which is the first year of significant runoff after three relatively dry years. Hence, it is not surprising that runoff during this year was somewhat lower than the trend considering the water needed to satisfy deficits built up during the dry years.

The contribution to runoff and ground water recharge from each landscape area is presented in Table 3. These data indicate that much of the runoff and recharge comes from the aspen area, which is a relatively small portion of the watershed. During most years, the sagebrush areas made no contribution to runoff and recharge, however only those years when these areas did contribute was there any appreciable runoff.

One would think that over the ten-year period, the

Table 4
Annual water balance summary (in mm) from aggregating the three landscape types

Year	Precip.[a]	ET[b]	Runoff	Δ*S* (soil)	Δ*S* (ground water)	Error	Error (%)[c]
1985	593	450	82	−31[b]	−32	123	21
1986	633	448	95	0[b]	14	77	12
1987	442	506	0	−65[b]	−30	30	7
1988	343	380	0	−37[b]	−22	22	6
1989	569	446	54	38[b]	8	23	4
1990	479	450	3	−21[b]	−2	48	10
1991	370	362	0	11	−28	25	7
1992	278	298	0	−18	−29	27	10
1993	694	413	61	100	75	46	7
1994	306	407	0	−118[b]	−19	37	12
Average	471	416	30	−14	−7	46	10

[a] Precipitation adjusted for drifting of snow.
[b] Indicates model-simulated values.
[c] Percentage of precipitation.

soil water storage term would become negligible, however the loss in soil water storage for the sagebrush areas was significant (Table 3). The low sagebrush lost an average of 15 mm per year and the mountain big sagebrush averaged a 24 mm loss. The combined loss for the soil and ground water averaged 21 mm annually, or a total of 210 mm for the period (Table 4). This large loss is due largely to a high precipitation year prior to the study, and relatively dry years near the end of the study. During water year 1984, the watershed received an effective precipitation of 829 mm, or approximately 175% of the average effective precipitation for the ten-year study period. The watershed, therefore, started the ten-year study period relatively wet. By contrast, precipitation for three of the four final years of the study ranged from 59 to 79% of average.

The annual water balance for each of the 10 years is given in Table 4. On average, the water balance error was 46 mm, or approximately 10% of the annual precipitation. The fact that all of the errors are positive indicates that estimated outflow from the watershed does not account for all of the estimated precipitation entering the watershed. Because potential errors in runoff and change in storage could not account for this magnitude of error over the study period, the likely causes are overestimation of effective precipitation or underestimation of ET and deep percolation.

A potential source of error in the effective precipitation and ET estimates arises from their spatial variability. However, differences in rainfall between the three gauges are not significant, and spatial distribution in snow accumulation is accounted for by using snow measurements at the grid locations to estimate drift factors. Estimation of the drift factors comes into scrutiny since these tend to elevate the effective precipitation above measured values and could introduce some error. Effective precipitation estimates based on the drift factors are somewhat higher than a simple average of the three precipitation gauges; annual basin-averaged effective precipitation is 471 mm versus an average of 454 mm from the three gauges. This is a difference of 17 mm which, at worst, is only a fraction of the average error. Therefore, it is unlikely that the drift factors and estimated effective precipitation is a major cause of the error.

Average leaf area index and ET vary relatively little within each vegetation zone. ET measurements reported by Artran (1996) for sites near I8 and I14 during the 12-day period of 1993 when both units were operating indicate that spatial variability between these two sites within the mountain big sagebrush area is not significant. Measurements between days 206 and 218 totaled 51 mm and were within 0.5% of each other. Although individual LAI transects at each location varied by 25% due to natural small scale variability in the vegetation, the maximum difference in leaf area index reported by Artan (1996) between these locations for similar dates was 0.15. Likewise, satellite-based soil adjusted

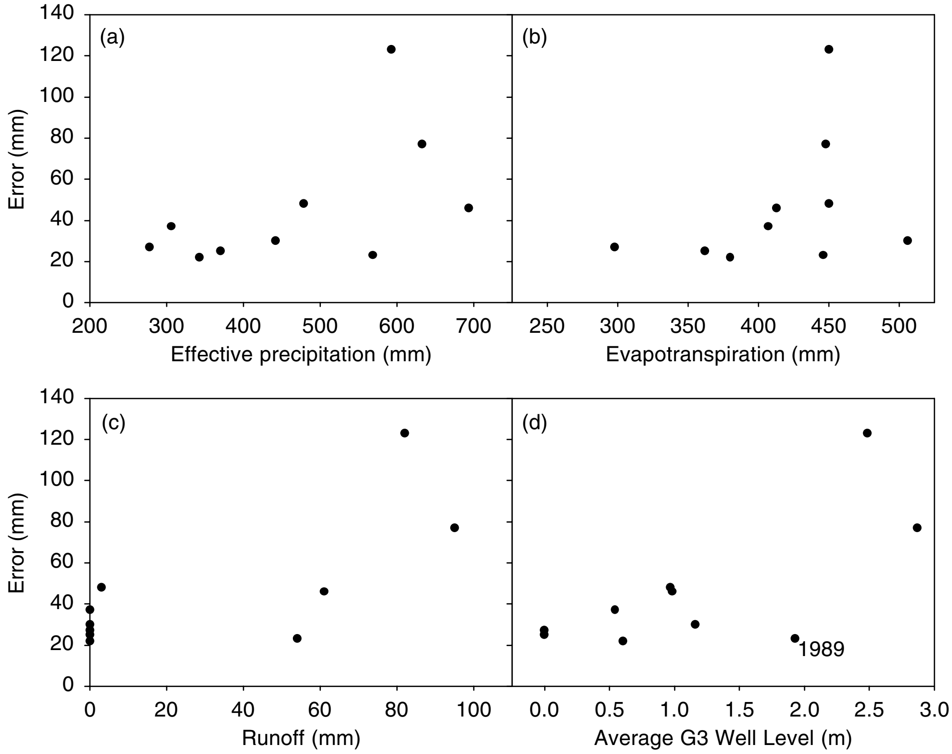

Fig. 5. Water balance error versus: (a) precipitation; (b) ET; (c) runoff; and (d) average ground water level in well G3.

vegetation index (SAVI) reported by Seyfried (1998) indicates very little variation between 30-m pixels within each vegetation zone. Increasing leaf area index for the three vegetation zones by up to 20% changed simulated ET by less than 3% over the ten-year simulation. This relatively small effect of leaf area is largely related to the timing and availability of water. Much of the precipitation input occurs as snowmelt when plants are not transpiring, and there is very little additional water available in the sagebrush areas for transpiration. Although simulated and measured values for ET in Table 2 may differ by typically 10% for short measurement periods, total simulated ET (815 mm) is within 2% of total measured ET (820 mm). Given the magnitude of average ET compared to effective precipitation for the low sagebrush and mountain big sagebrush (Table 3), it is unlikely that ET estimates can be off by more than 4%.

Admittedly a small percentage error in precipitation and ET could account for much of the water balance error. The 17 mm difference between measured and effective precipitation combined with the potential 4% error in ET could possibly account for 33 mm of the 46 mm average error. However, plots of the water balance error versus effective precipitation and ET have considerable scatter (Fig. 5a and b), which suggests that these are not likely a consistent source of the error. The 2 years with the largest error do have some of the highest effective precipitation values, which leads to further suspicion of the precipitation values. However, the year with the highest precipitation, 1993, has a rather modest error by comparison. Realizing that this year followed several dry years and would have lower than expected runoff, the linear trend for the remaining data in Fig. 4 above 450 mm of precipitation lends further support to the effective precipitation values for these wet years.

Inspection of Fig. 5c indicates a degree of order to the plot of water balance error versus runoff, particularly for annual runoff higher than 40 mm. It is unlikely for runoff to be the cause of the error, since runoff

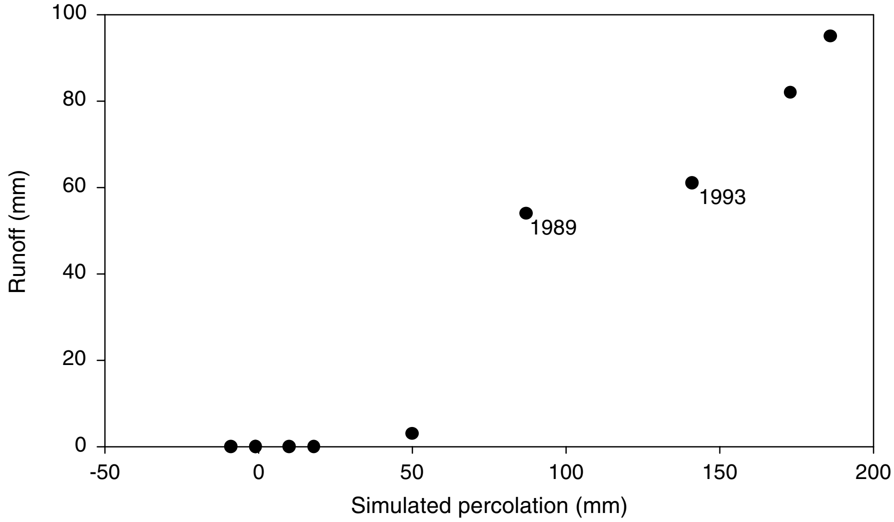

Fig. 6. Runoff versus simulated percolation beyond the root zone using the SHAW model.

measurements would need to be almost doubled to account for the water balance error. The plot of average ground water level at G3 versus the water balance error plotted in Fig. 5d shows a relation between amount of ground water and the water balance error ($r^2 = 0.51$). Piezometer G3 was selected for this analysis because it remains wet when most other piezometers have gone dry. With the exception of water year 1989, there is a noticeable trend between increasing error with increasing ground water level. Upon inspection of the changes in ground-water storage (Table 3), it is very likely that there is some loss to deep percolation through fractures in the bedrock underlying the watershed, as suspected by Flerchinger et al. (1998a). Even during years with no runoff, ground-water storage dropped by 20–30 mm (Table 4). Because the ground-water levels are in a fractured basalt layer approximately 20 m below the soil surface, it is unlikely that upward migration of water could account for this flow volume or that sagebrush roots could extract significant water from this depth. If either of these were the case, they would tend to be less of a factor during wetter years since ET demands could be satisfied by water stored within the soil profile. However the water balance errors are larger during years with higher ground water levels, which would provide more of a gradient for deep percolation out of the basin. Because deep percolation would tend to be more during wetter years, this could explain the trend toward larger errors during higher runoff years, as indicated in Fig. 5c.

On the basis of simulation results from the SHAW model, there is a strong relation between percolation beyond the root zone and measured runoff, as plotted in Fig. 6. Beyond a threshold value of 50 mm, the relation is quite linear; the slope of the line suggests that above this threshold, approximately 67% of the water percolating beyond the root zone produces runoff. Additional water is presumably lost to deep percolation through fractures in the dense basalt. The 1993 water year, which plotted as an outlier in the plot of runoff versus precipitation in Fig. 4, falls nicely in line with other years, suggesting that the model adequately accounted for precipitation required to satisfy soil water storage after the three dry years. Water year 1989, however, falls slightly off the linear trend for reasons unknown.

9. Summary and conclusions

The Upper Sheep Creek Watershed is a semi-arid, snow-fed rangeland watershed dominated by the processes of snowmelt, ET and subsurface water flow with ephemeral streamflow. Spatially variable precipitation, snow accumulation and vegetation present interesting challenges in this small, mountainous watershed. A ten-year water balance of the 26-ha Upper Sheep Creek Watershed was computed by disaggregating the watershed into landscape units (low sagebrush, mountain big sagebrush and aspen), computing a partial water balance for each and then aggregating these together to compute an overall water balance of the watershed.

Average annual effective precipitation for the watershed was 471 mm over the ten-year period. Spatial variability in effective precipitation has a large influence on hydrologic processes occurring in the watershed. This variability was found to increase with increasing precipitation. Runoff from the watershed averaged 30 mm and was correlated to effective precipitation above a critical threshold of approximately 450 mm necessary to generate runoff ($r^2 = 0.52$). ET averaged 416 mm, or nearly 90% of the annual precipitation, which is typical of semi-arid rangelands.

The average water balance error was 46 mm, or approximately 10% of the estimated effective precipitation for the ten-year period. The error is largely attributed to deep percolation losses through fractures in the basalt underlying the watershed. This more thorough analysis using additional continuous years of data confirmed suspicions of Flerchinger et al. (1998a) that the dense basalt layer is not as tight as initially assumed. Results suggest that losses to deep percolation are comparable to surface runoff from the watershed, which was difficult to ascertain with only 2 years of data used by Flerchinger et al. (1998a).

Simulated percolation of the water beyond the root zone correlated extremely well with measured runoff ($r^2 = 0.90$). Above a threshold of 50 mm, approximately 67% of the water percolating beyond the root zone produced runoff. This can have important ramifications in addressing subsurface flow and losses when applying a snowmelt runoff model to simulate runoff and hydrologic processes in the watershed.

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